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Seismic hazard of the western Makran subduction zone: insight from mechanical modelling and inferred frictional properties

Sepideh Pajang^{a,b,1}, Nadaya Cubas^b, Jean Letouzey^b, Laëtitia Le Pourhiet^b,
Seyedmohsen Seyedali^c, Marc Fournier^b, Philippe Agard^b, Mohammad Mahdi
Khatib^a, Mahmoudreza Heyhat^a, Mohammad Mokhtari^d

^a*Geoscience department, University of Birjand, Birjand, Iran*

^b*Institut des Sciences de la Terre Paris, ISTE_P UMR 7193, Sorbonne Université, CNRS-INSU,
75005 Paris, France*

^c*National Iranian Oil Company (NIOC), Tehran, Iran*

^d*International Institute of Earthquake Engineering and Seismology (IIEES), Tehran, Iran*

Abstract

Western Makran is one of the few subduction zones left with a largely unconstrained seismogenic potential. According to the sparse GPS stations, the subduction is accumulating some strain to be released during future earthquakes. To enhance the seismic hazard assessment, we here propose to study the finite deformation of the western Makran accretionary wedge. Mechanical modelling is used to retrieve the spatial variations of the frictional properties of the megathrust, and discuss its seismogenic potential. To do so, we first build a structural map along the Iranian part of the Oman Sea and investigate three N-S seismic profiles. The profiles are characterized by a long imbricated thrust zone that takes place at the front of the wedge. A diapiric zone of shallow origin lies in between the imbricated zone and the shore. Along the eastern and western shores, active listric normal faults seem to root down to the megathrust. Eastern and western domains have developed similar deformation, with three zones of active faulting: the normal faults on shore, thrusts ahead of the mud diapirs, and the frontal thrusts. On the contrary, no normal faults are identified along the central domain, where a seamount is entering into subduction. Two mechanical analyses are performed to retrieve the frictional prop-

Email address: sepideh.pajang@upmc.fr and sepideh.pajang@birjand.ac.ir (Sepideh Pajang)

erties of the megathrust. We first apply the critical taper theory to constrain the pore fluid pressure of the wedge. We then apply the limit analysis on two selected profiles. Along the eastern profile, a transition from very low to extremely low friction is required to activate the large coastal normal fault ($\mu_{deep}^{eff} = 0.01-0.06$, $\mu_{middle}^{eff} = 0.003-0.012$). To propagate the deformation to the front, an increase of friction along the imbricated zone is necessary ($\mu_{front}^{eff} = 0.017 - 0.031$). The method could not be applied on the incomplete western profile. However, since the deformation is similar to the eastern profile, the same transitions of friction are expected. The Central domain is also characterized by very low effective friction; but, the absence of normal fault does not allow to evidence any frictional transition.

Since dynamic effective friction coefficients are significantly lower than frictions at slow slip rate, the region of extremely low friction between the normal fault and the imbricated zone might reveal the location of a seismic asperity. The difference in deformation along strike would thus reveal the existence of two different asperities, one along the eastern domain and a second along the western domain. Since no earthquake have occurred in the region for, at least, the last 1000 years, an event of large magnitude may strike the Iranian Makran, in particular the Eastern domain.

Keywords: Makran, Seismic hazard, Frictional properties, Mechanical modelling

1. Introduction

The 1000 km long Makran is one of the few subduction zones left with a largely unconstrained seismogenic potential. The region indeed lacks historical records due to the sparsity of past settled population and is still poorly instrumented. However, the development of Chabahar and Gwadar Ports, and the risk of potential tsunamis for the region calls for a better hazard assessment. Along the Pakistani side, only three significant events in 1756, 1851 and 1945 have been reported [1]. The last one, with an estimated magnitude of Mw 8.1, is the only major event instrumentally recorded so far [1] (Figure 1a). Along the Iranian side, one major event dated back to 1483 could be linked to the subduction [2], but its location and date are strongly debated [3]. Recent geodetic and InSAR studies have shown that some

12 accumulation of elastic strain along the plate interface is required to explain the
13 velocity profiles on both the Pakistani and Iranian sides [4, 5, 6] (Figure 1b). All
14 studies proposed a relatively strong coupling along eastern Makran (where the 1945
15 earthquake occurred), a decrease along the central part, and a possible increase
16 westward. However, due to the sparsity of stations, large trench-coast distance, and
17 limited constraints on slab dip, these studies were unable to constrain neither the
18 along-dip nor the along-strike extent of possible coupled zones (e.g., [6]).

19 The Makran subduction zone is famous for its impressive ~ 7.5 km thick and 400
20 km wide accretionary prism, half of it sub-aerial [11]. Large accretionary prisms have
21 often been considered of limited seismic and tsunamigenic risk because of the rate-
22 strengthening nature of their unconsolidated sediments [12, 13]. This classical view,
23 the particular dimension of the Makran prism, and the lack of historical earthquake
24 led many authors to disregard the seismogenic potential of the area. However, the
25 idea of limited extent of earthquakes to accretionary prisms was strongly challenged
26 by the 2011 Mw 9.0 Tohoku-Oki and the 2004 Mw 9.0 Sumatra Andaman events
27 (e.g., [14]).

28 According to heat flow measurements and modelling (e.g., [15]), lab experiments
29 (e.g., [16]), drilling (e.g., [17]) or structural studies (e.g., [18, 19]), megathrust ef-
30 fective friction appears to differ significantly between aseismic or seismic areas.
31 Megathrusts undergoing mostly aseismic deformation are indeed characterized by
32 intermediate values of effective friction ($\mu \geq 0.1$), whereas lower effective (or ap-
33 parent) friction is found where major earthquakes occur ($\mu \sim 0.01-0.03$) [15, 18].
34 Such differences in effective friction are significant enough to produce wedge seg-
35 ments with varying morphologies and deformation patterns. For instance, a wedge
36 characterized by aseismic slip and thus intermediate friction along the megathrust
37 will reach more easily its critical state generating internal deformation (e.g., [18]).
38 On the contrary, seismic asperities characterized by very low effective friction coeffi-
39 cients are expected to construct a stable wedge, impeding any internal deformation
40 [18]. Moreover, the transition of effective friction coefficients from aseismic to seis-

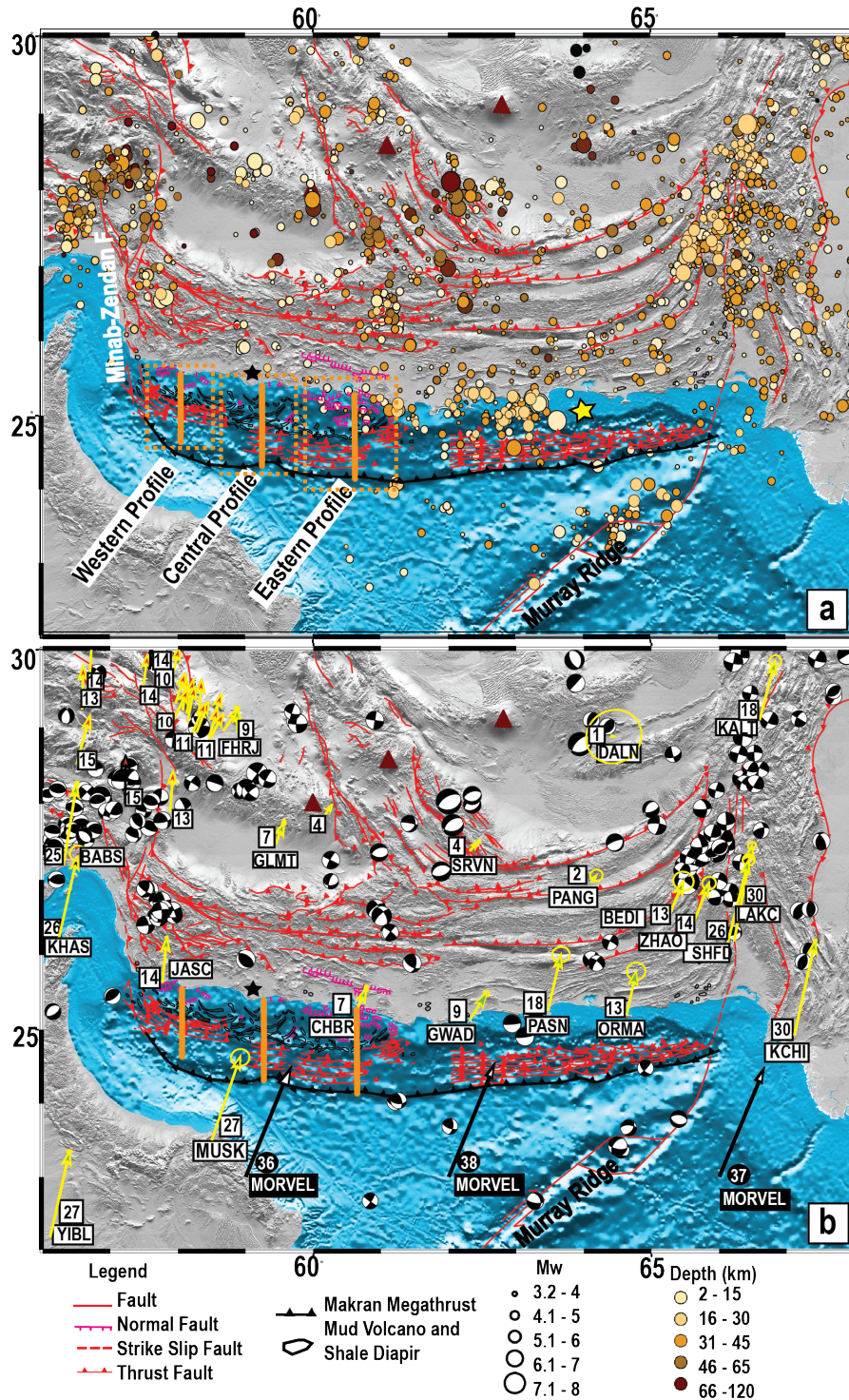


Figure 1: Seismo-tectonic map of Makran **a**. Seismicity, with earthquakes of magnitude ≥ 3.5 from 1945 to 2013 shown as colored circles. Red lines indicate known faults, mud volcanoes are shown by black contours. Positions of active mud volcanoes within Pakistani side from [7]. Imbricate thrust faults in the Pakistanis side from Smith et al. [8] picked from MCS data (multichannel seismic reflection) and correlated with bathymetric data. On the Iranian side, mud volcanoes, imbricate thrust faults, and normal faults are picked from PC2000 data released by NIOC. The onshore well in Iran is indicated by black star and the 1945 Mw 8.1 earthquake by yellow star. Thick orange lines indicate the 2D seismic lines used in this study. Based on the structural analysis western Makran can be separated in three different structural domains (Western, Central and Eastern) which are identified by dashed-orange boxes. **b**. Relative motions with respect to stable Eurasia. Yellow arrows are GPS velocity vectors with 95 per cent confidence ellipses [9, 5]; MORVEL velocities from DeMets et al. [10]. Focal mechanisms reported by gCMT.

mic asperities can also induce activation of splay faults [19]. As a consequence, the deformation observed in the continental wedge brings information on the mechanical behavior of the megathrust [20].

The Western Makran accretionary prism appears as an ideal candidate to study and link the finite deformation with spatial variations of megathrust frictional properties and seismic behavior. We here propose to take advantage of seismic profiles released by NIOC along the whole Iranian side to improve the assessment of seismic and tsunamigenic hazards of the region. We first present a brief seismo-tectonic description of the region before describing selected seismic lines. We then conduct a mechanical study to retrieve the spatial variation of the frictional properties of the megathrust. Seismic and tsunamigenic hazards are finally discussed in the last section.

2. Seismo-tectonic setting

The Makran subduction, which started during Early Cretaceous, is located in between two major collisional zones, Zagros to the West, and Himalaya to the East [21]. Transitions to collisional domains are accommodated by two major strike-slip systems, the Ornach Nal fault to the East, and the Minab-Zendan fault to the West (Figure 1). Current motions recalculated from seafloor spreading rates and fault azimuths for the major plates show comparable along strike convergence rates increasing slightly from 35.5 - 36.5 mm/a in western Makran to 39 mm/a in the East [10] (Figure 1b).

The Makran accretionary wedge is about 400 km wide, two-thirds of it being located onshore to the North. A narrow (few kilometers wide) coastal belt along which normal faults and mud volcanoes are prominent separates the on- and off-shore parts (e.g., [22, 21]). On its eastern side, the Makran accretionary prism accumulates up to 7.5 km of incoming sediments [11], brought by the Indus river and erosional portions of the growing prism. The incoming sediments consist of 3 km of Makran sands that are derived from the North [23]. They unconformably

69 overlie 4 km of Himalayan turbidites derived from the Indus fan to the East [11].
70 There are no major river system providing sediment input to the Oman Sea from
71 Iran. Hence, the amount of sediments decreases to the West. In the present deep-
72 water areas, sediments older than Lower to Middle Miocene are subducted together
73 with the oceanic crust, and the younger sediments above the décollement zone are
74 thrust and accreted [8].

75 Mud volcanoes have been reported both on- and off-shore along the Makran
76 accretionary wedge (e.g., [24, 7]). Mud volcanoes onshore appear to be associated
77 with E-W-trending fault zones [25] and sourced from the Upper Miocene Parkini
78 mudstones [26]. Active mud volcanoes are present and pictured on offshore seismic
79 profiles all along and near (within a few kilometers) the coast from West (e.g.,
80 [24]) to East, into Pakistan (e.g., [7]). The extruded very low-viscosity muds are
81 cold, with near-ambient temperatures with few degrees variation (e.g., [24]). Traces
82 of heavier hydrocarbons and isotopic compositions indicate gas generation from
83 thermally mature source rocks. Fossils collected in erupted mud suggest a Middle
84 Miocene over-pressured shale source [21].

85 Elevated marine terraces along the Makran coast indicate Quaternary surface
86 uplift [27]. The uplift rate derived from Uranium series on shell sand and ^{14}C
87 chronology, which yields minimum ages, has been estimated from 0 - 0.2 mm/a
88 (western) to 0.5 - 2 mm/a (eastern Makran) [27, 28]. Some studies have proposed
89 that the 2 m of coastal uplift during the Holocene along the Iranian side is associated
90 with a large ancient earthquake (e.g., [27, 2]). However, there is not enough evidence
91 to indicate that large earthquakes have occurred in the last 1000 yr.

92 The Makran region is not characterized by a high seismic activity (Figure 1a).
93 The eastern part of the subduction zone has been proposed as more seismically active
94 than the western segment (e.g., 1765, 1851, and 1945 earthquakes), which is in good
95 accordance with the occurrence of the 1945 Mw 8.1 earthquake [1]. According to
96 thermal modelling, eastern Makran may have a wide seismogenic zone [29]. In
97 contrast, only one major event was proposed for the western part, supposed to have

108 occurred in 1483, with an inferred magnitude of 8.0 [2, 1]. However, available oral
109 records for this event are few and they are unable to certify the location along the
100 subduction zone and the precise date. Musson [3] mentioned it was located in the
101 Strait of Hormuz and probably not on the subduction megathrust itself.

102 Few studies have tried to estimate the interseismic coupling [5, 6]. Frohling and
103 Szeliga [5] proposed a depth of locking of 38 km from the modelling of GPS velocities.
104 Penney et al. [6] tried to retrieve coupling with better constrained slab dips. This
105 study (with 17 stations among which 7 GPS stations along the coast distributed
106 over 1000 km long) was only able to propose partial coupling along two profiles,
107 one at 58°E, the second at 60.5°E. The authors found faster velocities along the
108 eastern profile. They admitted that while precise geometry of locking could not be
109 resolved with current data, the decrease of velocity towards the coast implied that
110 some elastic energy must accumulate along the wedge.

111 **3. Data acquisition and processing**

112 The study area is located offshore, with seismic profiles gathered under auspices
113 of NIOC (National Iranian Oil Company) in 2000 using conventional marine 2D seis-
114 mic methodology. These are part of the PC2000 project covering both the Persian
115 Gulf and the Oman Sea. The project comprises a seismic grid of 4x4 km in the west
116 and 8x8 km in the central and eastern domains (Suppl. Mat. **S1**). The grid covers
117 most of the seismic acquisition area, except in the deepest water areas towards the
118 abyssal plain to the South.

119 Two phases of acquisition were conducted by seismic vessel Bin Hai 517, with
120 basic acquisition parameters (recording length 8192 ms, fold of coverage 120, cable
121 length 6000 m,[30]). A last phase was recorded using seismic vessel Pejwak, with
122 basic acquisition parameters (recording length 20480 ms, fold of coverage (nominal
123 fold) 90, cable length 7200 m). Four lines were recorded in 20 seconds record length,
124 although only two of them extend across the prism and deformation front. We also
125 had access from NIOC to one onshore well located in Iran (black star, Figure 1).

126 All steps of the conventional processing work flow such as demultiplying, denoising,
127 and prestack time migration (PSTM) have been applied to available data. The
128 post-stack processing flow includes FX deconvolution and AGC (Automatic gain
129 control). We here present three of these NS profiles orthogonal to the trench, located
130 on the eastern, central and western part of the Iranian Makran. All seismic lines
131 were interpreted in order to generate a structural map and to identify different
132 structural provinces (Figure 1). In this study, the only source for depth conversion of
133 seismic sections are seismic velocities. For proper integration of seismic velocities in a
134 velocity model, we built a pseudo-3D velocity cube along each of our selected profiles
135 to account for structural complexities in the velocity model processing (Figure Suppl.
136 Mat. S1). The velocity model was obtained from the extrapolation of average
137 velocities derived from seismic processing velocity data using Dix equation [31].
138 Depth conversion using interval velocities has also been tested. However, due to
139 structural complexity, presence of shale diapirs and significant lateral variations
140 of velocities within the wedge [11], this method did not provide results in good
141 agreement with structures. Results obtained from average velocities for the three
142 selected profiles are discussed in the following sections.

143 4. Seismic data interpretation

144 4.1. Structural analysis

145 The Oman Sea Basin is characterized by a complex structure (i.e., several plate
146 boundaries, numerous thrusts and transfer faults [8, 23]) making the interpretation
147 of regional and deep horizons difficult. While several deep horizons can be inter-
148 preted within localized basins or sub-areas, their correlation into the next sub-basin
149 is uncertain since they may be separated by a tectonic high, mud volcanoes, transfer
150 faults and/or thrusts extending nearly to the seabed. Seismic horizons in the Oman
151 Sea Basin were thus defined herein by screening all seismic lines and picking out po-
152 tential correlations based on particularly strong reflectors, clear unconformities or
153 other characteristics as the cross-cutting BSR horizon and the décollement. Profiles

154 were further scrutinized whenever data quality allowed.

155

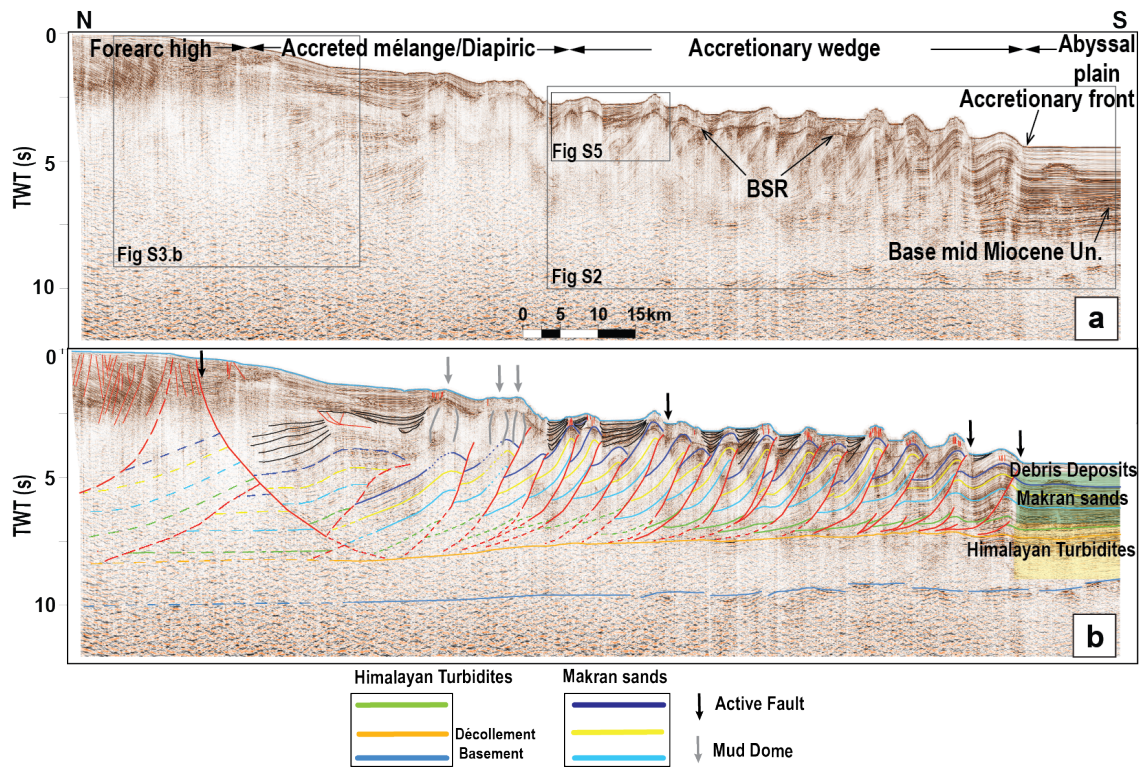


Figure 2: **a.** 2D offshore Eastern seismic profile across the Makran subduction zone from PC2000 in time (see Fig. 1 for location). **b.** Interpreted profile showing the link between the extensional province on the shelf and the contraction at the toe with thrust systems, as well as the main structural domains.

156 The wide continental wedge can be divided into three parts (Figures 1; 2; 4; 5):
157 (1) the accretionary wedge, (2) the accreted *mélange* and (3) the forearc high. In
158 western Makran, the accretionary wedge consists of a series of fold/thrust ridges
159 separated by small piggy-back basins. A Bottom Simulating Reflector is caused by
160 small amounts of free gas trapped below the sealing gas hydrate layer that cross-
161 cuts seismic reflectors. The horizon runs parallel to the seabed (Figures 2; 4) and
162 is observed on nearly all seismic lines, for water depths greater than 1300 m. The
163 accreted *mélange* is a complex structure, where nearly no primary bedding is pre-
164 served and no major structure is observed [32]. In general, the accreted *mélange* has
165 a diapiric character along the entire accretionary complex. Major seaward normal
166 faults that seem to root down to the *décollement* and shallower ones are observed
167 along the coast and the shelf on the eastern and western domains [33]. By cross-

168 cutting lithologies, their age is estimated to be younger than Late Miocene [34].

169

170 One onshore well is located on the Iranian side but does not intersect the seismic
171 profiles (black star, figure 1). However, the well shows that a large thickness of
172 young sediments is deposited close to shoreline, with a 4164 m drilled section and a
173 total depth in Upper Miocene sediments. This well consists of silty shales and shaly
174 sands, interbedded with thin sandstone intervals. Most of the published general
175 stratigraphy is available from onshore Pakistan [25, 33, 35]. As mentioned above,
176 the 7.5 km thick sediments [11] consist of about 4 km of Himalayan Turbidites which
177 have been interpreted as turbidites deposits and also as hemipelagic muds [33], and
178 are equivalent to the Hoshab and Panjgur formations of onshore Makran. The Late
179 Oligocene - Early Miocene Hoshab formation is composed of calcareous mudstones
180 with intercalation of sandstones [35] and the Middle Miocene Panjgur formation [36]
181 consists of thick bedded sandstone and conglomerates with thin bedded shale [33].
182 On top of the Himalayan Turbidites lies a 3 km unit of Makran Sands composed
183 of sandstone and siltstone of possible age Late Miocene to Pliocene [23]. This unit
184 is alike the Parkini and Talar formations of onshore Makran [26]. The Quaternary
185 hemipelagic sediments include a mixture of shelf and slope turbiditic sands, tur-
186 biditic silts or aeolian dust and, fluvial muds [37] that are equivalent to the Chatti
187 mudstone Ormara and Jivani formations [23] (Figure 2b).

188 *4.2. Eastern profile*

189 The Eastern profile, located south of Chabahar, is the longest available seismic
190 reflection profile, running over 140 km from the coast to the trench. The cross-section
191 was first drawn in time (Figure 2) and then converted to depth (Figure 3). Water
192 depth ranges from 0 m at the coastline to over 3300 m within the study area. There
193 are 6.5-7 km of sediments deposited on the abyssal plain, with a total thickness of
194 the sedimentary pile above the main basal décollement ranging between 7 and 11
195 km.

196 The proposed seismic interpretation builds on Grando and McClay [23]. The

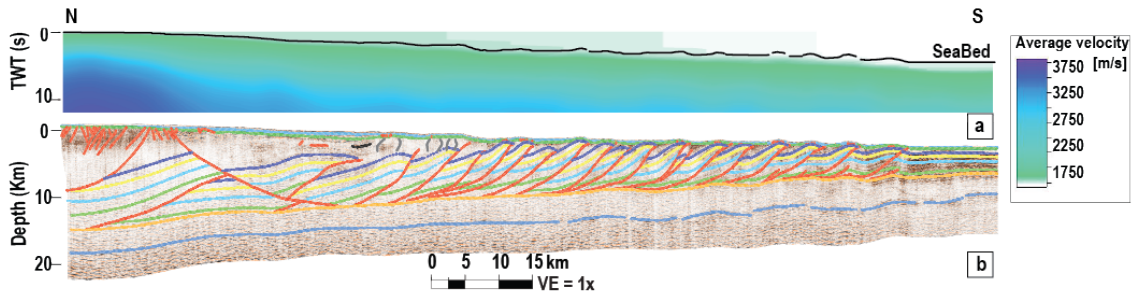


Figure 3: **a.** Smoothed average velocity profile for the Eastern profile obtained from the Seismic velocity data **b.** Depth section and interpretation (VE=1).

197 78 km long imbricated zone is composed of seventeen thrust sheets, with increasing
 198 spacing between the thrusts and decreasing dips towards the front [23] (Figures
 199 2, 3 and Suppl. Mat. S2). The mud diapir is located between a large seaward
 200 normal fault and the imbricated zone. Due to the weak resolution of the seismic
 201 image in this area, its source layer is debatable. According to our interpretation
 202 and previous studies [25, 24, 7, 21], it is most likely confined within a shallower
 203 and local décollement level. Due to the presence of shale and mud, the quality of
 204 the seismic data near the coast of the selected profile is quite poor. However, the
 205 comparison with surrounding lines allows us to root the seaward normal fault to
 206 the décollement (Figure Suppl. Mat. S3). The formation of these normal faults
 207 is probably related to the presence of deep duplexes [23, 36] visible on surrounding
 208 lines (Figure Suppl. Mat. S4), and to gravitational collapse [38]. From tilting of
 209 recent sediments, fanning, wedge shaped growth packages of sediment and fault tip
 210 at surface, we identify three main locations of active faulting: (1) the large seaward
 211 normal fault, (2) some thrust faults ahead of the diapiric zone, and (3) the frontal
 212 thrusts.

213 To validate our structural interpretation, we restored the cross-section according
 214 to Dahlstrom [39]’s method, based on the hypothesis that the length of beds remains
 215 constant during deformation in concentric folds. Cross-balancing was performed
 216 for the whole profile, excepted for the mud area where the volume conservation
 217 hypothesis does not hold (Figure Suppl. Mat. S6). The measured shortening from
 218 the deformation front to the northern tip of the section is of $65 \text{ km} \pm 10 \text{ km}$ ($32 \pm$

219 3 percent of the initial length).

220 *4.3. Central profile*

221 The total length of this cross-section is 122 km, excluding a 10 km gap located in
 222 the diapiric zone (Figure 4). Water depth ranges from 0 m at the coastline to over
 223 3100 m above the abyssal plain. A seamount of possible volcanic origin is visible
 224 at the front. The seamount is probably currently being subducted underneath the
 225 outermost part of the imbricated zone, generating strong uplift of the frontal folds.
 226 Thrusts on top of it are thus steeper in comparison with the Eastern profile. Only
 227 4 km of sediments is deposited on top of the seamount.

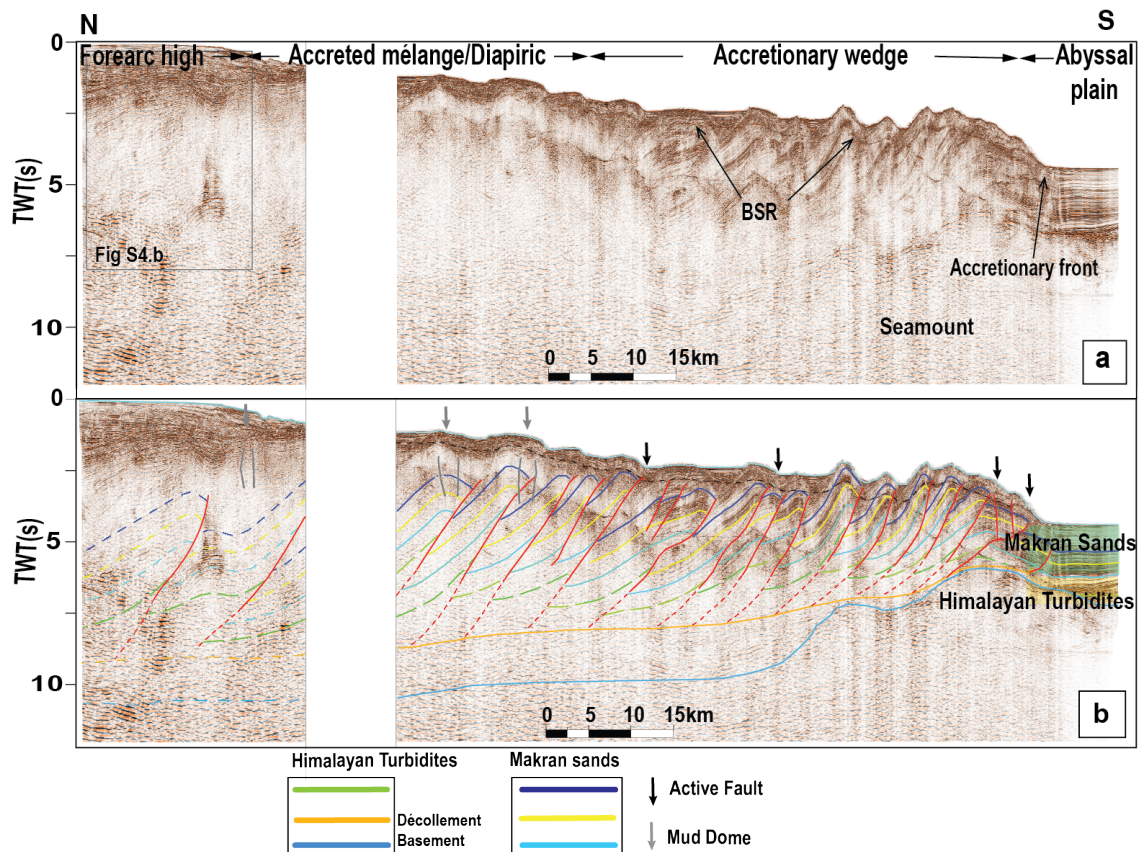


Figure 4: **a.** Central seismic profile of the study area (location on Figure 1). **b.** Interpreted profile. A seamount is visible at the accretionary front and has created antiformal stack structures.

228 Along this Central profile and the surrounding lines, we observe an increased
 229 number of shale diapirs with a larger spatial distribution compared to the eastern
 230 section (Figure 1). Moreover, unlike the eastern and western lines, we do not observe
 231 any evidence for major deep rooting normal faults (Figures Suppl. Mat. S3, S4).

232 The frontal thrusts and some thrusts located at the back of the imbricated zone are
233 identified as active.

234 Balancing of the cross-section suggests a total shortening, from the deformation
235 front to the diapiric zone (85 km of folded thrusts), of 50 ± 5 km (37 ± 2 percent
236 of the initial length), i.e., 5 percent longer than in eastern section. The difference is
237 most probably due to the presence of the seamount indentation causing a reentrant.

238 The seamount (Figure Suppl. Mat. **S7**) increases the basal slope and creates
239 an imbricate stack leading to an increased shortening of the shallow portion of the
240 wedge.

241 *4.4. Western profile*

242 The 97 km long Western profile does not reach the trench (Figure 5). Water
243 depth varies from 0 m to 1700 m to the southern end of the line. The profile is here
244 again characterized by a large seaward normal fault that seem to root down to the
245 décollement (Figure **S3**), a shale diapir zone, and the imbricated thrust zone. As
246 along the eastern profile, we identify three main active fault zones: the seaward
247 normal fault, thrusts ahead of the diapiric zone, and the frontal thrusts of the ac-
248 cretionary prism.

249

250 Seismic profiles allow proposing a structural map of the Iranian Oman sea with
251 the following first-order characteristics: a long imbricated thrust zone takes place
252 at the front of the wedge; a diapiric zone of shallow origin lies in between the
253 imbricated zone and the shore; some thrusts ahead of the diapiric zone are identified
254 as active. The eastern and western areas are characterized by active listric normal
255 faults located along the shorelines and rooting down to the décollement, whereas
256 the central domain shows no evidence of large normal faulting, a larger diapiric
257 zone, and is affected by a seamount entering into subduction. We thus propose to
258 separate western Makran in three different structural domains: Eastern, Central
259 and Western.

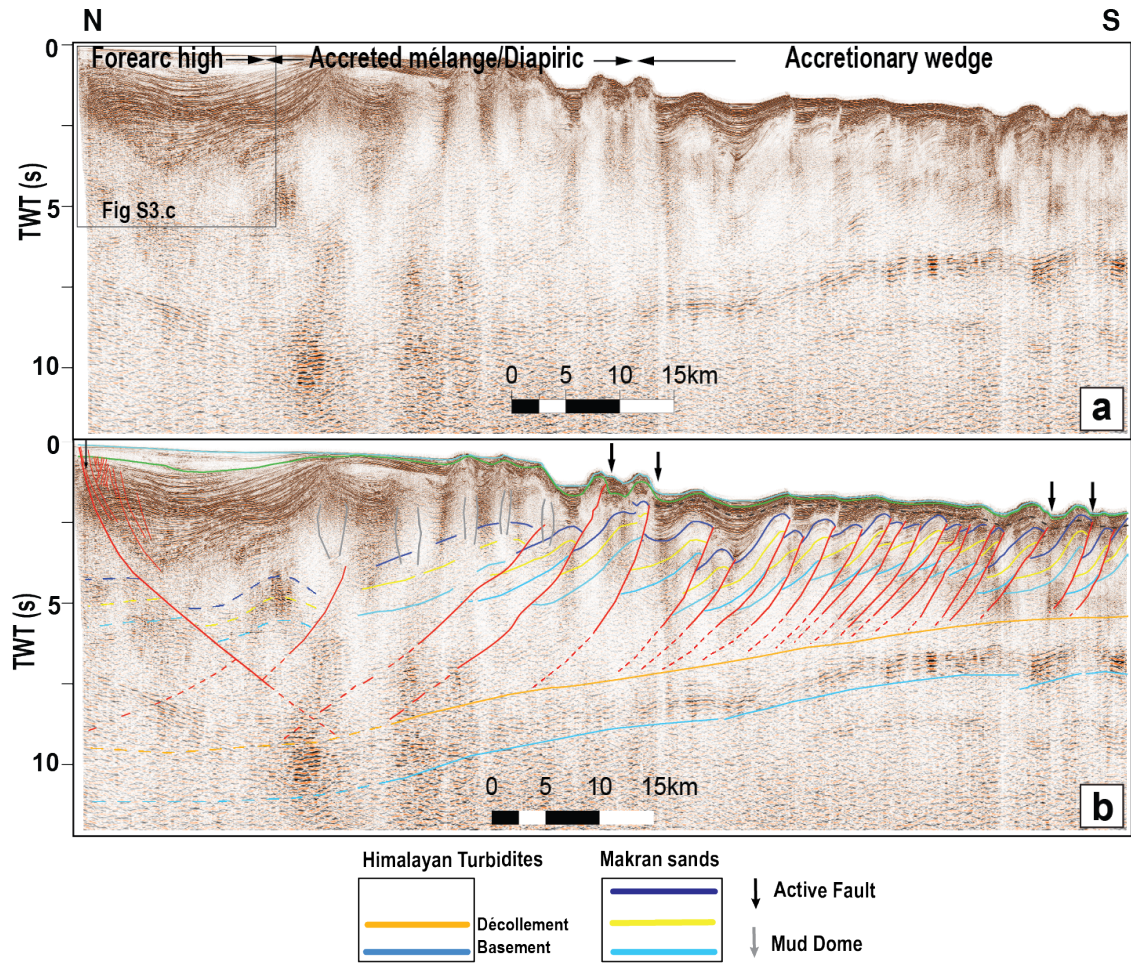


Figure 5: **a.** Regional N-S Western profile of the offshore Makran accretionary prism (location on Figure 1). **b.** Interpreted profile.

260 5. Mechanical modelling

261 In order to discuss the seismogenic potential of the megathrust, we propose
 262 to determine its frictional properties. Due to the high sedimentation rate, a high
 263 pore fluid pressure is expected in the sedimentary cover and needs to be estimated
 264 before running mechanical models. We thus conducted a two stage-analysis. First,
 265 we applied the critical taper theory over the entire area to confirm the location of
 266 active deformation, to constrain the pore fluid pressure of the wedge, and to provide
 267 a first estimate of the megathrust effective friction [40, 18]. Second, we applied the
 268 limit analysis approach, to get the along dip variation of the megathrust effective
 269 friction along the Eastern and Central profiles. Due to the seamount subduction,
 270 the décollement of the Central profile is quite rough, results are thus partial and

271 presented in supplementary material (Suppl. Mat. **S7**). Since the Western profile
272 does not reach the trench, the analysis could unfortunately not be carried out on
273 this profile.

274 5.1. Critical taper map

275 We first explore the mechanical state and the frictional properties of western
276 Makran with the help of the Critical Taper Theory (CTT) [41, 18]. This theory
277 allows us to relate the general shape of a wedge formed by its topographic slope
278 and the slab dip with the frictional properties of the wedge and megathrust. The
279 relation forms an envelope separating three different mechanical states (Figure 6a):
280 (1) a critical state, if the wedge follows the envelope. In that case, the megathrust
281 and the wedge are on the verge of failure, implying active faulting within the wedge;
282 (2) a stable state, if the wedge is inside the envelope. At stable state, the whole wedge
283 slides along its megathrust without producing any internal permanent deformation;
284 (3) unstable state, if the wedge is outside the envelope and cannot slide along its
285 megathrust. Frictional properties can only be estimated if the wedge is at critical
286 state and follows a critical envelop, which implies a co-variation of the topographic
287 slope and slab dip.

288 Following the method of Cubas et al. [18], we used ETOPO 1 for the bathy-
289 metric and topographic slope. The slab morphology could not be calculated from
290 the full seismic data set because of partial coverage. We thus used slab 2.0 [42],
291 after checking its consistency with the Eastern profile. We plotted cross-sections
292 perpendicular to the trench and constructed swath profiles to get the co-variation
293 of the topographic slope and slab dip with standard deviations for the inversion
294 procedure (Figure 6b). Along these profiles, we looked for segments at critical state,
295 i.e., parallel to a critical envelop (Figure 6a, b). For each of these segments, we
296 retrieved by inversion the friction of the wedge, the pore pressure ratio of the wedge
297 and the effective friction coefficient of the megathrust (Figure 6c, d). Effective fric-
298 tion coefficients are defined as $\mu^{eff} = \tan \phi^{eff} = (1 - \lambda) \tan \phi$, where λ is the
299 Hubbert-Rubbey fluid pressure ratio [40].

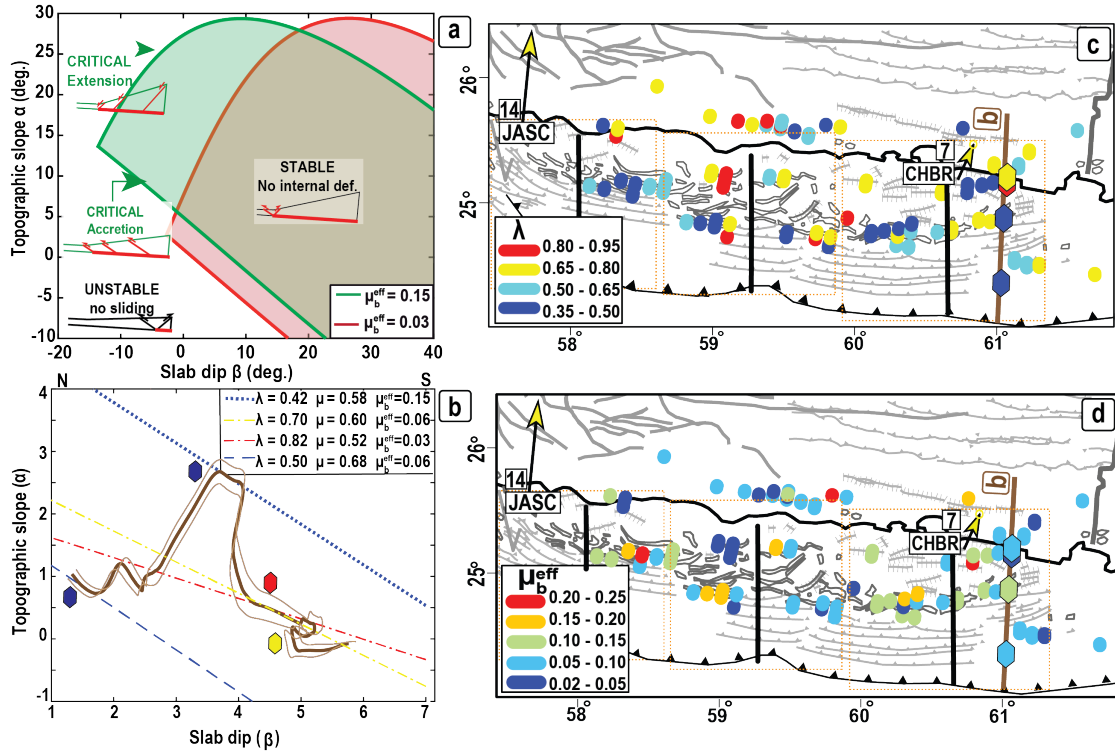


Figure 6: **a.** Critical envelopes and mechanical states for effective megathrust frictions of 0.15 and 0.03. **b.** Example of the α/β co-variation for a swath profile located along the eastern part of Makran (see location on c and d). Bold brown line: α/β values of the swath, cream lines: α/β mean \pm standard deviation. The diamond symbols denote the 4 segments identified at critical state (i.e., parallel to a critical envelope as described in a.). For each of these identified segments, the frictional properties are retrieved by inversion and given in the legend. The dashed color lines represent the lower branch of the critical envelop for each of the best fit set of parameters. Their location and properties are reported on panel c and d with diamonds. **c.** Pore pressure ratio of the wedge for segments at critical state and **d.** Effective friction coefficient of the megathrust for segments at critical state. The diamond symbols denote the four segments of b. The three structural domains are identified by dashed-orange boxes.

300 The slab morphology given by slab 2.0 is partly made of straight segments with
 301 kinks complicating the application of the method. However, although the errors on
 302 the slab geometry can affect the deduced effective frictions, the critical state usually
 303 remains valid for a large range of slab dip values [18].

304 Figure 6c, d shows patches of the prism at critical state, in particular in front
 305 of the mud diapirs, near the active out-of sequence thrusts identified on the seismic
 306 lines. Critical patches are also found in the Central domain where mud diapirs are
 307 present. Some critical patches lie along the coast consistent with the presence of
 308 deep duplexes. Active faulting identified on the seismic lines is thus supported by
 309 the CTT analysis. On the eastern and western borders, frontal and coastal critical
 310 patches get closer. Effective friction of the megathrust range between 0.017 and 0.23

311 and the pore fluid pressure in the wedge varies from hydrostatic to almost lithostatic.
312 Along the Eastern profile, we found a μ^{eff} of 0.1 in front of the mud diapirs, and
313 an internal pore fluid pressure ratio of $\lambda=0.6$.

314 Since this theory only provides effective friction of critical areas and since this
315 coefficient might be biased by the slab geometry, we then look for the frictional
316 properties allowing to reproduce the actual deformation with a complementary me-
317 chanical approach.

318 *5.2. Spatial variation of frictional properties from mechanical modelling*

319 To do so, we rely on the limit analysis method [43, 44], which is based on the
320 principle of virtual powers and the theorem of maximum rock strength [45]. In this
321 study, the Coulomb criterion is used for maximum rock strength. The principle
322 of virtual power consists in studying the collapse mechanisms, in our study faults,
323 by applying a virtual displacement over the structure. The method investigates
324 all possible distribution of faults as a function of the mechanical properties and
325 selects the optimal one minimizing the tectonic pushing force. Since deformation is
326 obtained from the chosen mechanical properties (such as frictions and cohesions), if
327 deformation is known, the frictional parameters can be retrieved by inversion. This
328 approach has already been applied to constrain megathrust friction [18, 19, 46].

329 For this study, we used the Optum-G2 software (Optum G2, 2013). To model the
330 Eastern profile, we used the geometry obtained from the depth conversion (Figure
331 3). Typical value of cohesions and internal frictions were chosen to reproduce the
332 stratigraphy, composed of the Makran Sands formed by interbeds with different
333 percentages of sandstone and siltstone, the Himalayan turbidites, a shale level for
334 the décollement, and a basaltic basement (Figure 7a) [47]. Since the Optum software
335 does not deal with over-pressure, we used effective friction coefficients, considering
336 a wedge pore fluid ratio of $\lambda = 0.6$ as inferred from the CTT.

337 To retrieve the effective friction coefficient along the megathrust, we searched for
338 the range of values reproducing the observed deformation, i.e., the seaward normal
339 fault rooting on the megathrust, the first thrusts located ahead of the mud diapirs

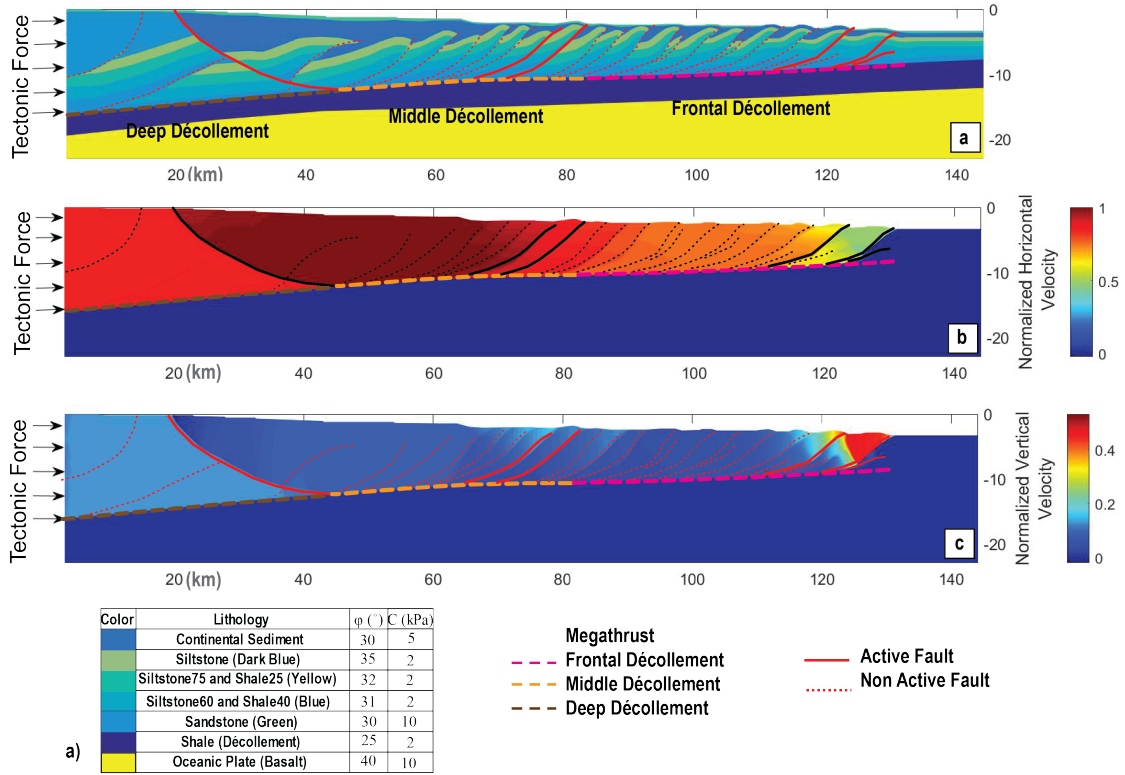


Figure 7: **a** Model set-up with three distinct décollement segments, to seek for the frictional properties reproducing the observed active deformation. The deep décollement has a dip of $\beta_{deep}=5^\circ$, and the pore fluid pressure ratio in the wedge is set to $\lambda=0.6$ as deduced from the CTT. **b** Normalized virtual vertical and **c** horizontal velocities showing the relative motion between the blocks, obtained for $\lambda = 0.6$, $\phi_{deep}^{eff} = 3^\circ$, $\phi_{middle}^{eff} = 0.6^\circ$ and $\phi_{front}^{eff} = 1^\circ$ ($\mu_{deep}^{eff} = 0.052$, $\mu_{middle}^{eff} = 0.01$ and $\mu_{front}^{eff} = 0.017$).

340 and the thrusts at the toe.

341 To activate the entire megathrust and fit the observed deformation, three seg-
 342 ments are needed, hereafter named frontal, middle and deep décollement (Figure 7).
 343 Although deformation might be sequential at very short-time scale, the displace-
 344 ment of the middle and frontal segments is generated by the displacement of the
 345 deepest segment. Hence, the deformation of one segment can not be modelled inde-
 346 pendently of the others. All possible effective frictions along these three segments
 347 were investigated, which led to seven different kinematics described in figure 8a.
 348 Kinematics **g** captures the observed active faults, and we now discuss the range of
 349 frictions obtained.

350 First of all, simulations indicate that a very low effective friction coefficient is
 351 required to activate the entire megathrust (Figure 8b). However, to activate the

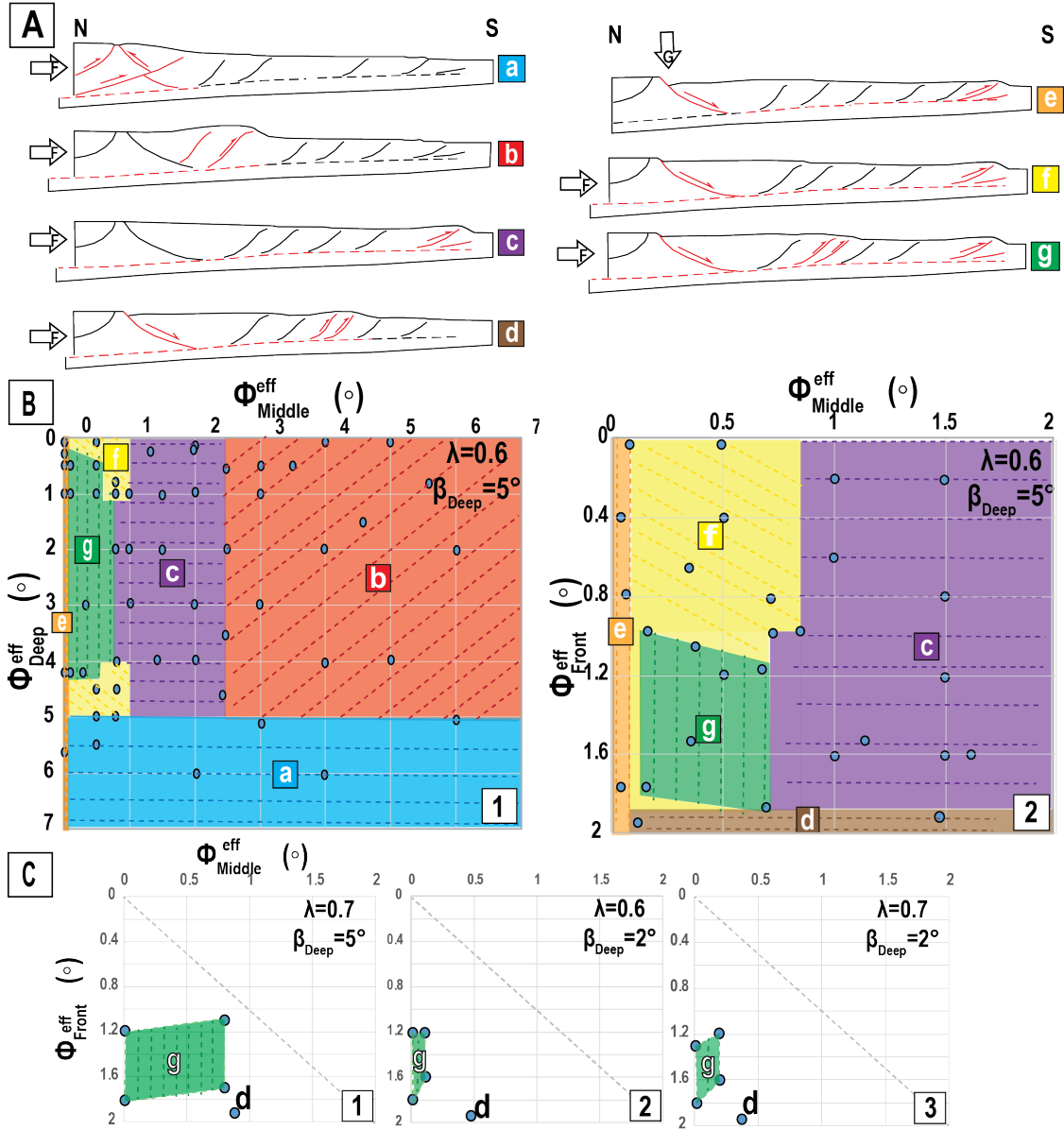


Figure 8: **A.** Kinematics obtained for the range of effective frictions along the décollement provided in figure **B** and **C**. **a-f** are failure models, **g** is the kinematics we seek to reproduce (**a**: Normal fault acts as reverse and décollement not active, **b**: Normal fault and décollement non-active, **c**: Normal fault and out-of-sequence thrust faults non-active, **d**: Décollement partially active, **e**: Gravitational normal fault, **f**: Normal faulting but non-active out-of-sequence thrust faults, **g**: Normal faulting, out-of-sequence thrust faults and full décollement active). **B.1** Exploration of kinematics for varying effective friction coefficients along the middle and deep décollement. Circles correspond to simulations, colored areas to kinematical domains, with kinematics **g** in green. **B.2** Exploration of kinematics for varying effective friction coefficients along the frontal and middle décollement. **C. 1-3.** Frictions along the frontal and middle décollement reproducing kinematics **g**. Two different pore pressure ratios ($\lambda=0.6$ and 0.7) and two different décollement dips ($\beta=2^\circ$ and 5°) are investigated.

352 normal fault, a decrease of friction from the deep segment to the middle one is
 353 necessary (Figures 8b, 7). To activate the out-of-sequence thrusts and propagate
 354 the deformation to the front, an increase of friction along the frontal segment is

355 necessary. As a consequence, the effective friction of the middle décollement has
 356 to be lower than the deep and frontal segments. Only a limited range of effective
 357 frictions can reproduce kinematics \mathbf{g} (Figure 8a and b, Table 1). For $\lambda = 0.6$, along
 358 the middle décollement, the effective friction can vary from $\mu_{Middle}^{eff} = 0.003$ to 0.012,
 359 with higher effective friction coefficients along the deeper and shallow segments as
 360 given in Table 1.

	Deep	Middle (lower bound)	Front
ϕ^{eff}	0.8 - 3.7°	0.02°	1 - 1.8°
μ^{eff}	0.01 - 0.06	0.003	0.017 - 0.031
	Deep	Middle (upper bound)	Front
ϕ^{eff}	1 - 3.7°	0.7°	1.2 - 1.9°
μ^{eff}	0.017 - 0.065	0.012	0.02 - 0.033

Table 1: Frictional properties along the megathrust reproducing kinematics \mathbf{g} with $\lambda = 0.6$.

361 To strengthen our analysis, we also tested different pore fluid pressure ratios
 362 and a lower slab dip ($\lambda = 0.7$, $\beta = 2^\circ$, Figure 8C) (a larger slab dip is inconsistent
 363 according to cross-balanced reconstruction). Again, the effective friction of the
 364 middle décollement needs to be lower than the deep and frontal segments.

365 This result is consistent with the critical taper theory. The middle segment, with
 366 a normal fault at the back and thrusts ahead should be on the verge of extensional
 367 critical limit, whereas the frontal segment with only thrust faults is at or close to
 368 the compressional critical state. For similar α and β , this change of state can only
 369 be achieved with lower effective friction along the middle segment.

370 The same procedure was applied to the Central profile (Figure 4). We again
 371 used the depth converted geometry (Figure Suppl. Mat. **S7**), and set all cohesions
 372 and frictions to standard values corresponding to the lithology (Figure Suppl. Mat.
 373 **S8**). We applied a pore fluid ratio $\lambda = 0.8$ as deduced from the CTT (Figure
 374 **6c**). Due to the presence of the seamount and the roughness of the frontal portion
 375 of the décollement, an extremely low effective friction is needed along the frontal
 376 décollement to propagate the deformation to the front ($\phi_{front}^{eff} = 0.4^\circ\text{-}0.6^\circ$, μ_{front}^{eff}
 377 $= 0.007\text{-}0.01$). However, the behavior of this frontal segment is probably closer to
 378 distributed deformation than to localised deformation along the megathrust [48],

379 and might not be captured with this approach. For the middle segment, we again
380 found very low effective friction coefficients, ranging from 0.1° to 2° ($\mu_{middle}^{eff} = 0.002$ -
381 0.034).

382 6. Discussion - Conclusion

383 According to the structural analysis, western Makran can be separated in three
384 different structural domains. The eastern one is characterized by active listric nor-
385 mal faults located along the shorelines and rooting down to the décollement, limited
386 extent of mud volcanoes, out-of-sequence thrusts ahead, and an imbricated zone.
387 The central domain shows no evidence of large normal faulting, presents a larger
388 diapiric zone, and the imbricated zone is affected by a seamount entering into sub-
389 duction at the front. The western domain is very similar to the eastern one. Three
390 zones of active deformation have been identified, supported by the CTT analysis:
391 listric normal faults on the shelf, several thrust faults ahead of the diapiric zone,
392 and the frontal thrusts.

393 To reproduce the deformation of the Eastern profile, an extremely low effective
394 friction is required between the seaward normal fault and the back of the imbricated
395 thrust zone. A decrease of effective friction is needed to activate the normal fault,
396 whereas an increase is required to propagate deformation to the trench. Along
397 the Central profile, an extremely low friction is also required along the middle
398 segment of the décollement. Although the method could not be applied to the
399 partial Western profile, the fact that deformation resembles that of the Eastern
400 profile suggests similar transitions of mechanical properties.

401 The very low and extremely low effective frictions are most probably explained
402 by near lithostatic pore fluid pressures. Low effective frictions have also been pro-
403 posed along the Pakistani side of Makran [49, 8], hence high pore pressures are
404 probably a common trend over the whole subduction zone due to the extremely
405 high sedimentation rate. However, these lithostatic pore fluid pressures can either
406 be a long-term or transient feature. Indeed, the inferred effective frictions are the

407 frictions necessary to reproduce the deformation. As a consequence, if deformation
408 is aseismic, these frictions would represent the effective friction at slow slip rate and
409 the pore fluid pressure would be a long-term feature. If deformation is acquired
410 during earthquakes or very shortly after, then the retrieved effective megathrust
411 friction would correspond to the dynamic friction. In that case, a already high pore
412 fluid pressure related to the sedimentation rate could reach even larger values due
413 to additional dehydration processes and thermal-pressurization acting during the
414 propagation of earthquakes [50, 51].

415 Considering that GPS data show some accumulation of elastic strain [4, 5, 6]
416 and that dynamic effective friction coefficients are lower than friction coefficients of
417 slow slip rate e.g., [15], this extremely low friction may reflect dynamic weakening
418 and the presence of a seismic asperity.

419 Splay faults branch upward from the plate boundary of subduction margins, and
420 are common in most subduction zones (e.g., [52, 53]). Their high tsunamigenic
421 potential due to their rupture during mega-earthquakes has been discussed in many
422 mega-tsunami events. A particular relationship between normal and splay faults
423 with seismic asperities has been observed during some recent major events. For
424 instance, during the 2010 Mw 8.7 Maule earthquake, a backthrust splay fault located
425 up-dip of the high slip patch has been activated leading to a significant uplift of the
426 Santa Maria island [53]. Two weeks after the main shock, two normal-faulting
427 aftershocks occurred along the down-dip limit of the high slip patch [54]. A seaward
428 normal fault rooting down to the megathrust was highlighted by crustal seismicity.
429 During the 2011 Mw 9.0 Tohoku-Oki earthquake, a landward normal fault was
430 activated at the transition between the high and the moderate slip patches [55].
431 Moreover, the Tohoku-Oki earthquake was followed by a shallow normal-faulting
432 earthquake sequence located along the Pacific coast, in a compressional domain [56].
433 An active crustal-scale normal fault system that dips landward and resembles the one
434 involved in the Tohoku-oki earthquake as also been identified along the Shumagin
435 Gap in Alaska, and has been associated to a high tsunamigenic risk [57]. The

436 activation of these normal faults has been related to variations of effective friction
 437 along the megathrust from seismogenic to aseismic portions of the megathrust [18,
 438 19] and to the release of gravitational potential energy [58]. According to McKenzie
 439 and Jackson [58], earthquakes would not only release elastic energy stored during
 440 the interseismic period, but also gravitational potential energy due to the sudden
 441 and strong decrease of the megathrust friction.

442 These splay fault examples thus support our interpretation: the extremely low
 443 friction between the normal fault and the out-of sequence thrusts of the imbricated
 444 zone would thus correspond to dynamic frictions reached during earthquake propa-
 445 gation, leading to the release of gravitational potential energy and the activation of
 446 the listric normal faults, and possibly of the out-of-sequence thrusts.

447 Therefore, the slip deficit, the deformation and frictions along the megathrust are
 448 all pointing to the presence of at least two seismic asperities located in between the
 449 listric normal faults and the out-of-sequence thrusts, one along the eastern domain
 450 and a second along the western domain (Figure 9).

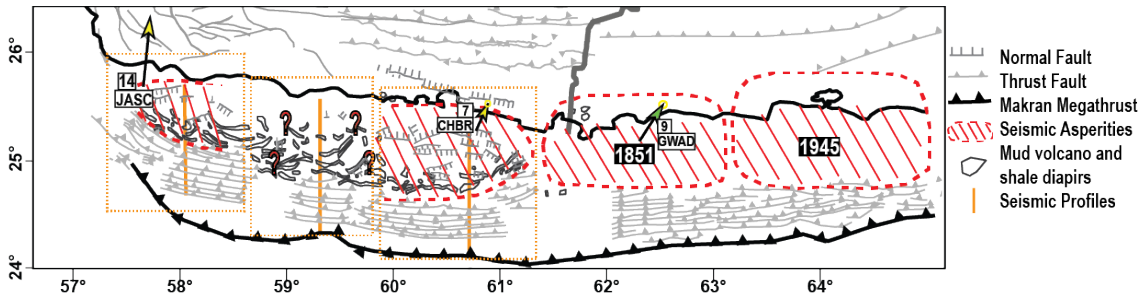


Figure 9: Areas of high seismogenic potential along the Makran subduction zone, according to our structural and mechanical analysis and past seismic ruptures from [4]. The three structural domains are identified by dotted-orange boxes

451 The behavior of the Central domain is more difficult to predict. The same low
 452 effective friction coefficient is consistent with a seismic asperity. However, frictional
 453 transitions are uncertain due to the lack of normal faults and more complex defor-
 454 mation at the front induced by the seamount. This absence can be interpreted in
 455 two ways: either the lack of normal faults is due to a change of slope, potentially
 456 induced by the seamount subduction at the front, or to the absence of a seismic

457 asperity. GPS stations along the central domain are thus mandatory for a better
458 assessment of seismic hazard in this region.

459 Several studies have also proposed that some continental wedge structures could
460 attest to the propagation of earthquakes to the surface as landward vergence thrusts
461 [20, 59]. Along western Makran, however, the prism is made of regular seaward
462 thrusts that do not show any evidence for such propagation. Tsunamigenic hazard
463 is thus probably limited to the identified normal faults and the out-of-sequence
464 thrusts [58].

465 Similar features along Pakistani Makran support our interpretation. Eastern
466 Makran is also characterized by a 70 km long imbricated zone made of regular sea-
467 ward thrusts [11, 49, 8], followed by a flat mid-slope terrace [49]. The frontal wedge
468 is characterized by a low taper suggesting some overpressure [49, 8]. The three last
469 major earthquakes, including the 1945 Mw 8.1, that struck Eastern Makran have a
470 similar along-dip extent as our proposed asperities: from the coastal region to the
471 rear of the imbricated zone at the change of slope. [1] (Figure 9). Based on 2D ther-
472 mal modelling and assuming uniform coupling between the 150° and 350° isotherms
473 depths, Smith et al. [29] have proposed a potentially wide seismogenic zone extend-
474 ing from the trench up to ~350 km (60 km depth). However, this along-dip extent
475 is not supported by InSAR observations, suggesting a high to moderate coupling
476 down to 20 km depth (180 km from the trench), coincident with the down-dip limit
477 of the 1945 earthquake, followed by a strong decrease with depth. The authors also
478 detected lateral variations of coupling revealing a possible along-strike segmentation
479 [4].

480 Therefore, along Western Makran where no large earthquake has been historically
481 reported despite the high convergence rate, a major event has to be expected. We
482 suggest that the magnitude of this event will depend on the mechanical behavior of
483 the Central domain, and the ability of the earthquake to propagate from the eastern
484 to the western asperity or to eastern Makran.

485

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489

490 **Appendix**

491 See supplementary material for supporting figures.

492

493 **References**

494 **References**

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